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Short Communication

More on the evolution of bed material waves in alluvial rivers

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Abstract

Sediment waves or pulses can form in rivers following variations in input from landslides, debris flows, and other sources. The question as to how rivers cope with such sediment inputs is of considerable practical interest. Experimental, numerical and field evidence assembled by the authors suggests that in mountain gravel-bed streams, such pulses show relatively little translation, instead mostly dispersing in place. This research has recently been the subject of discussion. In particular it has been suggested that (a) the equations of flow and sediment mass balance used in the analyses, and in most other morphodynamic analyses, require correction; (b) the dominance of dispersion appears only because the hyperbolic nature of the governing equations has not been adequately considered; and (c) the sediment transport equation used in the analyses does not lead to generalizable results. Here we suggest that (a) the relations for mass balance do not require the indicated correction; (b) the hyperbolic nature of the governing equations does not preclude the result of dispersion dominating translation in mountain streams; and (c) the general behaviour of an appropriate hyperbolic model of sediment waves (pulses) includes the relative roles of dispersion and translation, and is not affected by the precise choice of a sediment transport relation (as long as the choice is reasonable for the case in question). Copyright © 2005 John Wiley & Sons, Ltd.

Key words: dispersion theory; sediment waves; morphodynamic modelling

Introduction

Cao and Carling (2003) have provided a critique of the analysis of Lisle *et al*. (2001) on bed material waves (sediment waves, sediment pulses) in alluvial rivers. (In the present contribution the terms 'sediment wave' and 'sediment pulse' are used interchangeably.) This recent short communication (Cao and Carling, 2003) served three primary purposes: (a) to offer additional terms in the governing equations of river morphodynamics; (b) to demonstrate that the governing equations are universally hyperbolic; and (c) to point out limitations of the recent work of Lisle *et al*. (2001). Here we suggest that one of Cao and Carling's governing equations may not be correct, and the additional terms are likely to be negligible in most applications of interest, whether field or experimental. In addition, we suggest alternatives to several assertions in their short communication.

The 'Dispersion' Theory of Lisle *et al.*

Cao and Carling (2003) have labelled the theoretical development in Lisle *et al*. (2001) as a 'dispersion theory' of sediment-wave evolution. The formulation and solution given in Lisle *et al*. (2001), however, contain terms for translation as well as dispersion. Rather than ignoring translation, Lisle *et al*. (2001) conclude that conditions of bed-material transport that are common in gravel-bed channels lead to a diminishingly small value of the translation term relative to the dispersion term. Translation can, however, be significant in other circumstances. To illustrate this, Lisle *et al*. (2001) provide a field example of significant translation of sandy sediment waves in a gravel-bed channel with low Froude number, an observation that is consistent with their theory.

Lisle *et al*. (2001) provide a brief discussion of the numerical model of Cui *et al*. (2003a,b). Figure 13 of Cui *et al*. (2003b) provides an example of an adaptation of the numerical model for sediment waves (pulses) to the case of a low-slope, sand-bed stream. The same model that predicts minimal translation for gravel-bed mountain streams (using a sediment transport relation appropriate for that case) predicts substantial translation for low-slope sand-bed streams (using a sediment transport relation appropriate for that case). The difference in behaviour is not due to the use of a different sediment transport relation, however, but rather is due to the much lower Froude numbers prevailing in the latter case.

Correction to Equation 1 of Cao and Carling

The formulations of Cao and Carling (2003) for the conservation of total flow and sediment mass (equations 1 and 3 in their paper, which also appear in Cao *et al*. (2002) as equations 1 and 3, corresponding to Equations 1 and 2 below) contain additional terms that have not been included in the work of most previous authors.

$$\frac{\partial h}{\partial t} + U\frac{\partial h}{\partial x} + h\frac{\partial U}{\partial x} + \frac{\partial Y}{\partial t} = 0 \tag{1}$$

$$(1-p)\frac{\partial Y}{\partial t} + \frac{\partial}{\partial t}(q_s/U) + \frac{\partial q_s}{\partial x} = 0 \tag{2}$$

where t = time; x = streamwise distance; h = flow depth; U = cross-sectionally averaged streamwise velocity; Y = bed elevation; p = bed sediment porosity; q_s = volume sediment discharge per unit width ($q_s = ChU$, and C = average volume sediment concentration).

The last term on the left-hand side of Equation 1 and the second term on the left-hand side of Equation 2 are the terms added by Cao and Carling (2003); they argue for the significance of these terms in morphodynamic modelling. In this section we suggest that Equation 1 may not be correct, and then provide our proposal for a replacement. We further illustrate in the next section that the additional terms in the corrected equations are likely to be negligible in most rivers.

To elaborate on Equation 1, the conservation equation for the combined mass of water and sediment is first considered. This conservation equation takes the form

$$\frac{\partial(h+Y)}{\partial t} + \frac{\partial(q_w + q_s)}{\partial x} = 0 \tag{3}$$

in which q_w = Uh denotes the water discharge per unit width. Equation 3 can be rewritten as

$$\frac{\partial h}{\partial t} + U\frac{\partial h}{\partial x} + h\frac{\partial U}{\partial x} + \frac{\partial Y}{\partial t} + \frac{\partial q_s}{\partial x} = 0 \tag{4}$$

Comparing Equations 1 and 4, it is found that the only condition under which Equation 1 is correct is the case of vanishing $\partial q_s/\partial x$, a condition that is not satisfied in general. We suggest that the correct replacement for Equation 1 is obtained by substituting Equation 2 into Equation 4 so as to eliminate the term $\partial q_s/\partial x$, resulting in Equation 5:

$$\frac{\partial h}{\partial t} + U\frac{\partial h}{\partial x} + h\frac{\partial U}{\partial x} + p\frac{\partial Y}{\partial t} - \frac{\partial}{\partial t}(q_s/U) = 0 \tag{5}$$

Why the Extra Terms in Cao and Carling can be Strictly Neglected for Most Cases of Interest

The corrected version of flow and sediment mass conservation is Equation 5; Cao and Carling's (2003) equation for sediment conservation (Equation 2) remains unaltered. Equation 2 contains the term $\partial(q_s/U)/\partial t$, and Equation 5 contains the terms $-\partial(q_s/U)/\partial t$ and $p\partial Y/\partial t$ that have not been included in most previous morphodynamic analyses. There is a specific reason why they have not been included, as demonstrated below.

Equations 5 and 2 are first placed in dimensionless form using parameters characterizing a reference equilibrium state (and denoted with overbars):

$$\tilde{Y} = \frac{Y}{\bar{h}}, \quad \tilde{h} = \frac{h}{\bar{h}}, \quad \tilde{U} = \frac{U}{\bar{U}}, \quad \tilde{q}_w = \frac{q_w}{\bar{q}_w} = \frac{q_w}{\bar{U}\bar{h}} \tag{6a,b,c,d}$$

$$\tilde{t} = \frac{t}{\bar{h}/\bar{U}}, \quad \tilde{q}_s = \frac{q_s}{\bar{q}_s}, \quad \tilde{x} = \frac{x}{\bar{h}} \tag{6e,f,g}$$

where the variables with the tilde are the dimensionless variables; $\bar{h}$ denotes equilibrium water depth; $\bar{U}$ denotes cross-sectionally averaged velocity at the equilibrium condition; $\bar{q}_w = \bar{U}\bar{h}$ denotes the equilibrium water discharge per unit width; and $\bar{q}_s$ denotes the equilibrium volume sediment transport rate per unit width. Substituting Equations 6a–g into Equations 5 and 2, the following respective forms are found:

$$\frac{\partial \tilde{h}}{\partial \tilde{t}} + \tilde{U}\frac{\partial \tilde{h}}{\partial \tilde{x}} + \tilde{h}\frac{\partial \tilde{U}}{\partial \tilde{x}} + p\frac{\partial \tilde{Y}}{\partial \tilde{t}} - \varepsilon\frac{\partial}{\partial \tilde{t}}(\tilde{q}_s/\tilde{U}) = 0 \tag{7a}$$

$$(1-p)\frac{\partial \tilde{Y}}{\partial \tilde{t}} + \varepsilon\frac{\partial}{\partial \tilde{t}}(\tilde{q}_s/\tilde{U}) + \varepsilon\frac{\partial \tilde{q}_s}{\partial \tilde{x}} = 0 \tag{7b}$$

where

$$\varepsilon = \frac{\bar{q}_s}{\bar{U}\bar{h}} = \frac{\bar{q}_s}{\bar{q}_w} = \bar{C} \tag{8}$$

and $\bar{C}$ denotes the equilibrium value of C.

Most rivers transport far less sediment than water, so that the condition $\varepsilon << 1$ prevails even during floods (e.g. Parker, 1976). A clear illustration of this is provided by Mulder and Syvitski (1995). When rivers meet the sea, they normally form surface plumes because even sediment-laden river water is usually much lighter than sea water. In order for a river to plunge and form a bottom turbidity current when it meets the sea, the density of the river water must exceed that of sea water, i.e. near 1·026 tons/m^3. Assuming a specific gravity of sediment of 2·65, in order to plunge the concentration C of sediment in a river must thus exceed the rather small value of 0·0158. Mulder and Syvitski (1995) document that while such streams do exist (e.g. the Yellow River, China), they are rare. The great majority of streams have much lower values of C, even during floods.

The non-dimensionalization of time of Equation 6e gives the ratio of time t divided by a characteristic hydraulic time scale $\bar{h}/\bar{U}$. Over time scales characteristic of hydraulic response, then, $t \sim \bar{h}/\bar{U}$ and thus $\tilde{t} \sim 1$. Thus under the condition $\varepsilon << 1$ Equation 7b approximates to the form

$$\frac{\partial \tilde{Y}}{\partial \tilde{t}} = 0 \tag{9}$$

Equation 7a likewise approximates with the aid of Equation 9 to the form

$$\frac{\partial \tilde{h}}{\partial \tilde{t}} + \tilde{U}\frac{\partial \tilde{h}}{\partial \tilde{x}} + \tilde{h}\frac{\partial \tilde{U}}{\partial \tilde{x}} = 0 \tag{10}$$

That is, over hydraulic time scales bed elevation can be taken to be unchanging in time, and flow mass balance takes the form corresponding to a fixed-bed channel.

A morphodynamic time scale t^* associated with the erosion and deposition of sediment can be defined as

$$t^* = \varepsilon\tilde{t} \tag{11}$$

That is, as t^* changes by an order-one amount, $\tilde{t}$ must change by order $1/\varepsilon$, i.e. a very large amount compared to the characteristic dimensionless hydraulic time scale of unity. Substituting Equation 11 into Equation 7b, which governs morphodynamic change, the following equation is obtained:

$$\varepsilon(1-p)\frac{\partial \tilde{Y}}{\partial t^*} + \varepsilon^2 \frac{\partial}{\partial t^*}(\tilde{q}_s/\tilde{U}) + \varepsilon \frac{\partial \tilde{q}_s}{\partial \tilde{x}} = 0 \tag{12}$$

Dropping the second order term in ε, the equation approximates to the following form

$$(1-p)\frac{\partial \tilde{Y}}{\partial t^*} + \frac{\partial \tilde{q}_s}{\partial \tilde{x}} = 0 \tag{13}$$

Transforming Equations 10 and 13 back to the original dimensioned variables, it is found that the appropriate forms for Equations 5 and 2 for a morphodynamic analysis are

$$\frac{\partial h}{\partial t} + U\frac{\partial h}{\partial x} + h\frac{\partial U}{\partial x} = 0 \tag{14a}$$

$$(1-p)\frac{\partial Y}{\partial t} + \frac{\partial q_s}{\partial x} = 0 \tag{14b}$$

That is, the extra terms in Cao and Carling (2003) are negligible in most rivers.

The extra terms in Equations 5 and 2 are not universally negligible. Some rivers, such as the Yellow River, China, can transport sediment in such high concentrations that the basis for the quasi-steady approximation breaks down (i.e. the assumption that $\varepsilon << 1$ is not valid any more). This is also the case for lahars, or heavily sediment-laden river flows resulting from volcanic eruptions. Any treatment of debris flows, as well as the highly sediment-laden flow that sometimes occurs immediately downstream of a debris flow after it deposits, can be expected to require the extra terms. Cao and Carling (2003) have done the research community a service by pointing this out.

The Hyperbolic Nature of the Governing Equations

Most of the communication in Cao and Carling (2003) is devoted to a demonstration that the governing equations are universally hyperbolic. The authors suggest that the hyperbolic nature of the governing equations has been studied thoroughly and is well known (e.g. de Vries, 1965; Gill, 1988; Lai, 1991; Sloff, 1993; Cui *et al*., 1995; Cui and Parker, 1997b). In particular, Cui and Parker (1997b) demonstrate that the evolution of sediment waves can be dominated by dispersion despite the hyperbolic nature of the governing equations.

Types of Sediment Waves and Sediment Bores

Cao and Carling (2003) suggest that there are two types of sediment wave: one that can become progressively flatter with diminishing amplitude and one that can evolve into a shock wave with a sharp front (shock or discontinuity). Ribberink and van der Sande (1985) and Gill (1988), for example, discuss the propagation of a leading-edge discontinuity associated with an impulsive increase in sediment feed. Sharp discontinuities also take the form of the delta front (sediment bore) as a river progrades into standing water (e.g. Needham and Hey, 1991).

Sediment waves with discontinuities have not been discussed in the sediment wave (sediment pulse) work of the group of researchers who have authored this short communication (e.g. Lisle *et al*., 1997, 2001; Cui *et al*., 2003a,b; Cui and Parker, in press). The phenomenon, however, can be simulated with our family of models by setting the appropriate upstream and/or downstream boundary conditions (e.g. Cui *et al*., in press a). Simulated sediment waves with sharp fronts can be found in Cui *et al*. (1996), Cui and Parker (1997a), Cui and Parker (1999), and Cui *et al*. (in press a). In particular, Cui *et al*. (1996) applied the traditional shock-capturing method, and Cui and Parker (1997a) applied a shock-fitting method and successfully simulated the progress of the discontinuous sediment front in addition to the downstream fining associated with bed aggradation; Cui *et al*. (in press a) simulated one of the runs previously simulated by Cui *et al*. (1996) and Cui and Parker (1997a) with the sediment wave model of Cui *et al*. (2003b); and Cui and Parker (1999) were able to apply the same algorithm to simulate the progress of the front of the sediment deposit resulting from the disposal of mine waste sediment into the Ok Tedi–Fly River system in Papua New Guinea. Examples of the simulations of Cui *et al*. (1996), Cui and Parker (1997a), and Cui and Parker (1999) are shown in Figures 1 and 2. Details of the simulations are not presented here; readers are referred to the original references. The common features of the formation of sharp front in Figures 1 and 2 are a very low Froude number downstream of the

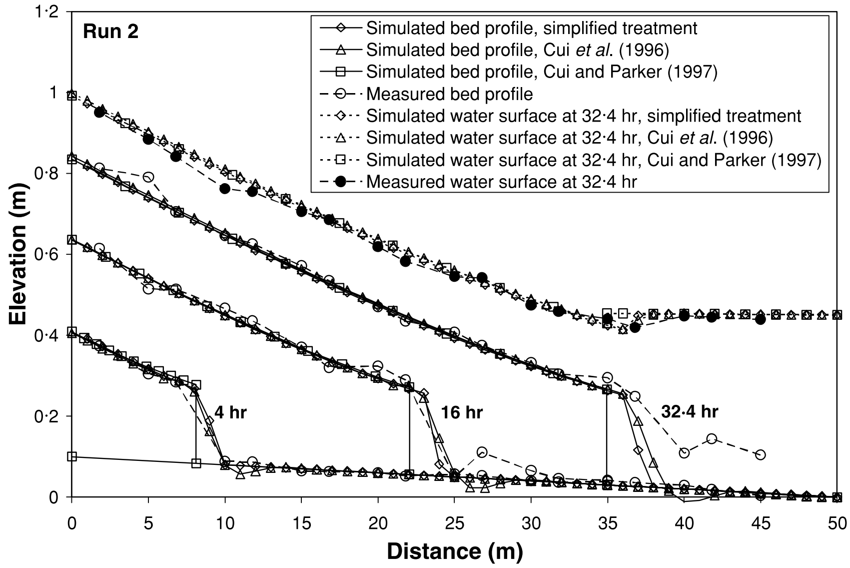

Figure 1. Run 2 of the St. Anthony Falls Laboratory downstream fining experiment (Paola *et al.*, 1992), simulated by Cui *et al.* (1996), Cui and Parker (1997a) and Cui *et al.* (in press a). The diagram is from Cui *et al.* (in press a), and the results labelled 'simplified treatment' are simulated with the model of Cui *et al.* (2003b). Flow is from left to right.

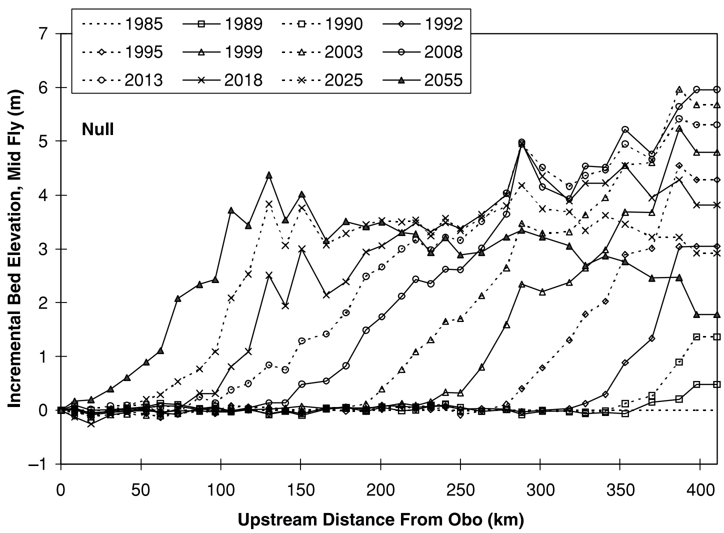

Figure 2. Simulated sediment deposition in the middle Fly River, Papua New Guinea, for mine operation continued until year 2010 without sediment remediation (Cui and Parker, 1999). Flow is from right to left.

front, and a sediment load far exceeding the equilibrium sediment transport capacity that continues either throughout the run (experiments) or was sustained for many years (field).

It is useful to point out that the evolution of the sediment front in Cui *et al*. (1996) and Cui and Parker (1997a) is very similar to that observed in the experiment performed by Needham and Hey (1991). Needham and Hey (1991) suggest that a sediment front such as those shown in Figures 1 and 2 advances with constant speed. In addition, they imply that the conservation equations for flow, momentum and sediment cannot admit solutions where there are discontinuities such as sediment fronts (sediment bores). We suggest that neither statement is accurate. Standard techniques such as shock fitting and shock capturing provide excellent means for describing discontinuities in hyperbolic systems. Cui *et al.* (1996) have used shock capturing, and Cui and Parker (1997a) have used shock fitting to describe the migration of deltaic fronts. The migration speed of the front is predicted as a product of the analysis. For example, consider an equilibrium condition under constant inflow rates of water and sediment. A sediment front can be created by significantly increasing the sediment feed rate while maintaining the same water discharge. As long as the water discharge and the increased sediment feed rate remain constants, the migration speed of the front must gradually decline in time, due to the steadily increasing rate of extraction of sediment consumed in aggrading the bed upstream of the front. In the case of relatively short experiments, the migration rate may appear to be constant, but only as a local approximation.

The application of a shock-capturing method seeks a 'weak solution' and requires the introduction of artificial viscous terms in the governing equations. In general, this technique produces acceptable results when the shock is pronounced, such as in the case of Cui *et al.* (1996) shown in Figure 1. When the shock, or discontinuity, becomes small, however, the artificial viscosity often leads to artificial waves on the channel bed that are of the same order of magnitude as the true disturbance itself, so giving unacceptable results. Because of this shortcoming, the shock-capturing method has not been used in the sediment wave (pulse) modelling by the authors. Instead a simplified method was adopted, i.e. flow parameters were calculated with the standard backwater formulation for low Froude number, and the quasi-normal assumption was used for high Froude number. This simplified technique produces results that agree well with more complicated methods such as those used in Cui *et al.* (1996) and Cui and Parker (1997a), as demonstrated in Cui *et al.* (in press a), and shown here as Figure 1.

The propagation of a sediment front is always associated with an increase in slope compared to that prevailing before the passage of the front, as correctly pointed out in Needham and Hey (1991). This feature appears to be missing from figure 1b of Cao and Carling (2003).

The Sediment Transport Relation of Meyer-Peter and Muller

Cao and Carling (2003) state that the recent dispersion/diffusion models (Lisle *et al.*, 1997, 2001) entirely hinge upon particular sediment transport function(s), and thus can only be applicable to limited situations. They do not, however, provide any comparative results to validate their statement that particular bed load equations lead to contrasting predictions of relative rates of dispersion and translation under the same conditions.

It is a fact that Lisle *et al.* (1997) used the relation of Meyer-Peter and Muller (1948) as part of their model. There is, however, nothing peculiar about this relation, and substituting it with any other reasonable sediment transport relation will not compromise the basic characteristics of a numerical model of sediment waves (pulses). For example, in subsequent work (Sutherland *et al.*, 2002; Cui *et al.*, 2003b; Cui and Parker, in press; also summarized in Lisle *et al.*, 2001) the surface-based bedload equation for gravel mixtures due to Parker (1990) was used. The numerical model successfully reproduced the evolution of a sediment wave (pulse) due to a landslide in the Navarro River, California, without any adjustment in parameters. In addition, Cui and Parker (1999, in press) and Cui *et al.* (2003b) used Brownlie's bed material equation (Brownlie, 1982) in a model of sediment waves (pulses) and obtained reasonable results. These same simulations indicate that as opposed to mountain streams, sediment pulses in low-slope sand-bed streams can show significant translation as well as dispersion.

The State of the Art in Sediment Wave Modelling

Cao and Carling (2003) state that the 'quality of mathematical river modelling remains uncertain primarily because there are no universal sediment functions or hydraulic resistance relations for the underlying physics of interaction between water flow and sediment transport'. Although we agree with Cao and Carling (2003) that there remain uncertainties in numerical modelling, we are optimistic that sediment transport modelling can be a useful tool for understanding the general behaviour of channel morphology and for solving practical problems, provided that the models are applied at the appropriate scale and interpreted with caution. For example, our family of numerical models on sediment waves or pulses (e.g. Lisle *et al.*, 1997; Cui *et al.*, 2003b; Cui and Parker, in press) has helped us to understand the relative importance of different parameters (e.g. Froude number, initial wave length, initial wave amplitude, grain size distribution, particle abrasion) in the general behaviour of sediment waves; and application of the Cui and Parker (in press) model to dam-removal projects and the disposal of waste sediment from mines has helped stakeholders evaluate management alternatives (Cui and Wilcox, in press; Cui *et al.*, in press a,b; Cui and Parker, 1999).

Conclusion

We suggest that the additional terms in the modified governing equations in Cao and Carling (2003) either require amendment or are usually unnecessary for most morphodynamic modelling, including that of the evolution of sediment waves (pulses). The hyperbolic nature of the governing equations is well understood and does not contradict the dominance of dispersion in gravel-bed rivers. Sediment waves with an advancing front have been studied extensively prior to the sediment wave work of Lisle *et al*. (2001) by various researchers (e.g. Cui *et al*., 1996; Cui and Parker, 1997a, 1999). The sediment transport relation of Meyer-Peter and Muller (1948) used in Lisle *et al*. (1997) can be replaced with a different sediment transport equation without compromising the general characteristics of a model of sediment waves (pulses). The family of sediment wave (pulse) models of the group who authored this short communication (Lisle *et al*., 1997; Cui *et al*., 2003b, Cui and Parker, in press) has provided insight into the relative importance of parameters and the general behaviour of sediment waves. The models have also been adapted and used successfully to solve engineering problems such as dam removal and the disposal of mine-derived sediment (Cui and Wilcox, in press; Cui *et al*., in press a,b, Cui and Parker, 1999).

Cao and Carling (2003) may have misunderstood certain aspects of Lisle *et al*. (2001) because several related publications would not have been available to them. The complete corpus of publications is introduced here. All papers that are under review or in press but not yet published can be downloaded at http://www.stillwatersci.com/sedimenttransportpubs.htm.

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